PO/SO PHASES: PROPAGATION VELOCITY ACROSS A 1,500-KM-LONG, DEEP OCEAN HYDROPHONE ARRAY

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(Received September 30, 1986; Revised May 18, 1987)

Po/So phases from numerous earthquakes along the margin of the Northwestern Pacific Basin were successfully recorded by a 1,500-km-long ocean bottom hydrophone (OBH) array deployed for two months near Wake Island. Data from ten shallow-focus (<100 km) events at 20° (2,200 km) to 34° (3,800 km) epicentral distance were used to compute propagation velocities across the array for the onsets of Po and So. The values found are 7.96 ± 0.05 km/s for Po and 4.58 ± 0.06 km/s for So. Travel-time intercepts associated with these velocities are, respectively, -7.21 ± 2.40 s and -12.84 ± 7.61 s. These velocities are significantly lower than those previously reported for the Northwestern Pacific using single station data. Although the low velocities and negative intercepts could simply be the result of bias in the data, they might also be due to the effects of azimuthal anisotropy along travel paths, or to increasing propagation velocity with lithospheric age, or to a different mode of propagation near the source.

1. Introduction

Late in the summer of 1981, the Hawaii Institute of Geophysics (HIG) deployed a 1,500-km-long linear array of twelve ocean bottom hydrophones (OBH's) near Wake Island (Fig. 1). Half of the instruments began recording on 12 August and ended on 23 September. The remaining half started on 3 September and ended on 15 October. The total recording time was about 65 days, and all twelve instruments were in operation from 3 September through 23 September. Of the twelve OBH's deployed, ten were recovered, and nine of these had quality data throughout their operational period. Monitored concurrently were three bottom hydrophones of the Wake Island Array (WIA), a 40-km array of hydrophones located near, and cabled directly to, Wake Island. WIA data were used in the study of first arrivals and were considered to be part of the OBH array data. All of the OBH and WIA instruments were in ocean depths between 5,265 m and 5,657 m (Table 1). Of more than 130 events identified from seismic phases in the recordings, 108 were located by the National Earthquake Information Service (NEIS; Table 2

Fig. 1. Map showing OBH locations and epicenters of earthquakes recorded by the array (see Tables 1 and 2). Instruments operating from 12 August through 23 September are indicated by open circles. Those operating from 3 September through 15 October are indicated by closed circles. Three instruments that did not work properly are flagged. Event numbers (from Table 2) are indicated for epicenters of events specifically used in this study. Bottom hydrophones of WIA are at the approximate position of OBH 4.

Table 1. Hydrophone coordinates and depths.

Hydrophone	Latitude (N)	Longitude (E)	Water depth (m)
OBH 1	23.18	163.81	5,657
OBH 2	22.10	164.80	5,265
OBH 3	21.28	165.54	5,600
OBH 4	20.23	166.47	5,495
OBH 7	17.25	169.03	5,445
OBH 8	16.41	169.72	5,450
OBH 10	14.52	171.28	5,635
OBH 11	13.37	172.19	5,600
OBH 12	12.55	172.79	5,580
WIA 74	20.60	166.50	5,400

Table 2. Events recorded by the OBH array (Data from NEIS Monthly Lists).

No.	Date	Time	Location	<i>h</i>	<i>m</i>	<i>M</i>	<i>n</i>	<i>Po</i>	<i>So</i>
1	Aug. 12	22:35	Solomon Is.	33	4.9	—	3	—	—
2	13	02:57	Tonga Is.	191	5.4	—	3	—	—
3	14	05:42	Volcano Is.	33	4.7	—	3	—	—
4	14	06:24	Molucca Passage	38	5.5	5.3	1	—	—
5	14	09:05	Honshu, Japan	58	—	—	2	—	—
6	15	10:30	Alaska	53	5.1	—	2	—	—
7	15	19:53	Philippine Is.	152	4.8	—	1	—	—
8	16	23:54	Kuril Is.	33	5.6	5.0	3	—	—
9	17	02:17	West Irian	34	5.7	5.8	3	—	—
10	17	17:07	Fiji Is.	383	5.5	—	2	—	—
11	18	05:29	Banda Sea	34	5.1	4.8	1	—	—
12	19	01:41	Kermadec Is.	185	5.1	—	1	—	—
13	19	03:01	Fiji Is.	507	4.8	—	1	—	—
14	19	06:06	Loyalty Is.	25	5.6	4.9	1	—	—
15	20	02:19	Santa Cruz Is.	71	5.0	—	3	—	—
16	20	12:53	Mariana Is.	193	3.8	—	4	—	—
17	20	15:10	Kermadec Is.	347	4.9	—	1	—	—
18	21	14:29	Bonin Is.	509	4.6	—	1	—	—
19	23	01:59	Loyalty Is.	100	5.8	—	1	—	—
20	23	12:00	Kuril Is.	40	6.0	5.8	4	3	3
21	24	15:46	Aleutian Is.	56	5.2	—	2	—	—
22	25	06:56	Honshu, Japan	325	4.8	—	4	—	—
23	25	07:16	Tonga Is.	33	5.9	5.7	1	—	—
24	25	07:22	Tonga Is.	433	5.7	—	1	—	—
25	25	20:07	Mariana Is.	33	4.9	4.4	1	—	—
26	26	04:51	Mariana Is.	40	5.2	5.1	5	—	—
27	26	16:32	New Britain	74	5.7	—	6	—	—
28	26	18:56	Honshu, Japan	230	4.4	—	3	—	—
29	28	09:04	Alaska	71	5.1	—	3	—	—
30	30	11:36	Fiji Is.	609	5.4	—	6	—	—
31	31	06:14	Komandorsky Is.	33	4.7	3.8	1	—	—
32	Sept. 1	07:23	Tonga Is.	33	5.8	5.7	4	—	—
33	1	09:29	Samoa Is.	25	7.0	7.7	6	—	—
34	1	18:38	Tonga Is.	33	5.7	5.3	3	—	—
35	1	23:55	Tonga Is.	33	5.6	5.4	3	—	—
36	2	08:44	Samoa Is.	33	5.3	5.5	1	—	—
37	2	09:24	Honshu, Japan	58	5.5	—	2	—	—
38	3	03:59	Hokkaido, Japan	52	4.7	—	4	—	—
39	3	04:29	Philippine Is.	93	5.8	—	4	—	—
40	3	05:35	Kuril Is.	45	6.6	6.6	10	6	6
41	3	19:39	Honshu, Japan	44	5.6	5.4	4	—	—
42	3	19:44	New Guinea	136	4.8	—	5	—	—
43	4	11:15	Philippine Is.	645	6.0	—	10	—	—
44	4	23:44	Solomon Is.	38	5.4	5.3	9	—	—
45	6	11:02	Loyalty Is.	31	5.9	6.2	1	—	—
46	7	15:11	Fiji Is.	231	5.2	—	3	—	—
47	7	16:20	Honshu, Japan	440	4.9	—	3	—	—
48	7	19:06	Honshu, Japan	33	5.8	5.5	7	6	5
49	7	20:07	Honshu, Japan	29	5.1	4.7	7	5	4

Table 2. (continued)

No.	Date	Time	Location	<i>h</i>	<i>m</i>	<i>M</i>	<i>n</i>	<i>Po</i>	<i>So</i>	
50	Sept.	8	19:26	Kurils Is.	46	5.7	5.4	5	—	3
51		10	03:29	Solomon Is.	110	4.9	—	3	—	—
52		10	23:21	Mariana Is.	13	5.6	5.2	10	5	—
53		11	08:33	Fiji Is.	554	5.2	—	5	—	—
54		12	03:40	Fiji Is.	302	5.2	—	3	—	—
55		12	07:15	Kashmir	33	6.2	5.9	3	—	—
56		12	14:51	Hokkaido, Japan	111	4.9	—	7	—	—
57		12	16:14	Bonin Is.	33	4.6	—	2	—	—
58		13	01:20	Honshu, Japan	39	4.8	4.8	4	—	—
59		13	02:17	Eastern Kazakh	0	6.0	4.5	4	—	—
60		13	20:24	Honshu, Japan	87	4.9	—	3	—	—
61		14	15:08	Honshu, Japan	33	5.4	5.0	4	4	—
62		15	14:12	Banda Sea	102	5.9	—	2	—	—
63		15	20:43	Taiwan	167	4.9	—	3	—	—
64		17	06:19	Banda Sea	33	5.7	5.8	1	—	—
65		17	08:23	Loyalty Is.	30	5.7	6.6	2	—	—
66		17	12:42	Fiji Is.	356	5.2	—	10	—	—
67		17	21:12	Kamchatka	33	4.9	3.9	2	—	—
68		19	07:27	China	561	4.4	—	3	—	—
69		20	04:39	Honshu, Japan	33	4.4	—	2	—	—
70		22	06:55	Mariana Is.	33	4.2	—	6	—	—
71		24	17:20	Bonin Is.	33	5.7	5.3	5	3	2
72		25	03:25	New Guinea	116	4.8	—	1	—	—
73		25	10:21	Bonin Is.	33	4.4	—	2	—	—
74		25	14:30	Kermadec Is.	45	5.9	5.9	5	—	—
75		25	15:01	Honshu, Japan	25	5.5	6.1	2	—	—
76		28	03:36	Honshu, Japan	31	5.5	5.3	5	—	—
77		28	17:56	Kermadec Is.	323	6.0	—	5	—	—
78		29	23:02	Tonga Is.	226	5.1	—	5	—	—
79		30	07:04	New Guinea	123	5.4	—	4	—	—
80		30	23:03	Pacific Ocean	10	5.9	5.2	2	—	—
81		30	23:37	Hokkaido, Japan	52	4.9	4.9	2	—	—
82	Oct.	1	12:14	Novaya Zemlya	0	5.9	3.8	4	—	—
83		1	13:10	New Ireland	85	5.0	—	2	—	—
84		1	16:02	Kermadec Is.	33	5.6	5.1	5	—	—
85		1	17:04	Kuril Is.	33	5.9	5.7	5	4	4
86		1	19:00	S. Nevada	0	4.9	—	2	—	—
87		2	05:51	Savu Sea	109	4.6	—	4	—	—
88		2	15:13	Mariana Is.	113	5.0	—	5	—	—
89		3	07:21	Kuril Is.	75	5.1	—	4	—	—
90		3	08:26	Hokkaido, Japan	62	4.4	—	2	—	—
91		3	16:48	Philippine Is.	238	5.0	—	2	—	—
92		4	00:01	New Guinea	33	5.9	6.3	4	—	—
93		4	04:11	Honshu, Japan	38	5.2	5.0	4	—	2
94		4	10:18	Solomon Is.	58	5.1	—	2	—	—
95		4	10:27	Solomon Is.	24	5.7	5.4	5	—	—
96		5	04:28	China	534	4.7	—	2	—	—
97		6	07:40	Aleutian Is.	23	5.2	4.3	3	—	—
98		7	03:02	Fiji Is.	620	5.8	—	5	—	—

Table 2. (continued)

No.	Date	Time	Location	<i>h</i>	<i>m</i>	<i>M</i>	<i>n</i>	<i>Po</i>	<i>So</i>
99	Oct. 7	08:23	Solomon Is.	41	5.8	5.3	5	—	—
100	7	15:03	Samoa Is.	33	4.8	—	2	—	—
101	7	17:48	Philippine Is.	595	5.3	—	4	—	—
102	9	12:19	Solomon Is.	50	6.0	6.4	5	—	—
103	9	19:46	Mariana Is.	96	4.7	—	5	—	—
104	11	00:36	Minahassa Pen.	94	5.6	—	4	—	—
105	13	15:53	Kamchatka	112	5.3	—	3	—	—
106	14	12:29	Mariana Is.	205	5.0	—	5	—	—
107	14	20:09	Mariana Is.	131	4.9	—	5	—	—
108	15	01:47	Honshu, Japan	47	6.0	5.4	5	—	—

h, Event depth (km); *m*, body-wave magnitude; *M*, surface-wave magnitude; *n*, number of instruments in array which recorded body-wave phases P, Po, or So from this event; *Po*, number of Po arrivals from this event used in first-arrival study; *So*, number of So arrivals from this event used in first-arrival study.

and Fig. 1). The primary purpose of the experiment was to acquire data of critical importance in understanding a phenomenon known as high-frequency Pn/Sn, long-range Pn/Sn, or Ph.f./Sh.f.; but referred to here as Po/So or Ocean P/Ocean S after WALKER (1982).

Po/So phases were first observed in the North Atlantic and have been found throughout the North, Western, and Central Pacific. In the Western Pacific, the seismic velocities and high-frequency character of these phases have been reported by ASADA and SHIMAMURA (1976), SHIMAMURA and ASADA (1976), SHIMAMURA *et al.* (1977), WALKER (1977a, b), WALKER *et al.* (1978), SUTTON *et al.* (1978), OUCHI (1981), OUCHI *et al.* (1983), NAGUMO *et al.* (1981), WALKER *et al.* (1983), BUTLER (1986), and CESSARO and DUENNEBIER (1987). First-arriving Po/So energy travels with a fairly constant velocity of about 8.0 and 4.6 km/s, respectively, and peak amplitude arrivals have velocities of about 7.6 and 4.5 km/s, respectively, which are comparable to basal crustal rates. At distances of about 18° ($\approx 2,000$ km), observed frequencies of Po and So are as high as 30 and 35 Hz, respectively; and at distances of about 30° ($\approx 3,300$ km), as high as 15 and 20 Hz, respectively. The signal/noise ratios for Po/So phases are often much greater than the ratios of corresponding normal, mantle-refracted P and S phases; and in many instances no P's or S's can be found despite the presence of very strong Po's and So's. Except for the SOFAR channel of the world's oceans, the Po/So waveguide appears to be the earth's most efficient acoustical waveguide. Also, it seems probable that the phenomenon is a dominant feature of all of the world's oceans and marginal seas—with the entire lithosphere, or large portions of it, serving as the Po/So waveguide.

Recent efforts have been made, using synthetics, to determine the mechanism of Po/So propagation (e.g., FUCHS and SCHULZ, 1976; STEPHENS and ISACKS, 1977; MENKE and RICHARDS, 1980; SUTTON and HARVEY, 1981; GETTRUST and FRAZER,

1981; SERENO and ORCUTT, 1985; MALICK and FRAZER, 1986), although many essential characteristics of these phases are still poorly known. Data collected before the present experiment, composed almost entirely of events recorded at single stations and over an unevenly distributed range of epicentral distances, made difficult the precise determination of the travel-time curve between 0° and 40° , as well as the attenuation of these phases as a function of frequency and distance. The high frequencies and large signal/noise ratios observed would indicate that Po/So attenuation must be significantly smaller than that of normal, mantle-refracted P or S phases at similar distances. When a frequency-independent model for Q was fit to some older Po/So data with frequencies between about 2 and 10 Hz, the resulting Q values were greater than 5,000 in most cases (WALKER *et al.*, 1978; SUTTON *et al.*, 1978). These values are much higher than those generally found for upper mantle travel paths from P and S data at frequencies below 2 or 3 Hz. However, the high frequencies and large signal/noise ratios might also be due in part to a frequency-dependent Q combined with low noise levels at the higher frequencies (greater than 4 Hz) where Po/So are most prominent. SUTTON *et al.* (1983) found two methods for extracting a relationship between Q and frequency from single-station Po/So data. They found that Q was approximately proportional to frequency over the range 2–9 Hz and that the Q of So was higher than the Q of Po for any given frequency. Unfortunately, the older data (i.e., different earthquakes recorded at only single stations) make it hard to separate attenuation from other factors that influence the observed amplitudes such as magnitude, focal depth, epicentral distance, focal mechanism, and possibly azimuth.

To eliminate some of these uncertainties, an experiment was designed to record Po/So across a linear, 1,500-km-long array consisting of 12 OBH's. The data provided by such an experiment would permit determination of Po and So velocities independent of the source parameters, and thus, hopefully, would eliminate a major source of scatter and error in velocity estimates. Also, data from the experiment would facilitate a nearly direct measure of attenuation as a function of frequency and travel time (ignoring near station effects, only a spreading term would have to be assumed in order to determine the apparent Q).

The array was aimed toward northern Japan and the southernmost islands of the Kuril chain (Fig. 1). This target was chosen because of its long history of moderate-to-large earthquakes and several successful recordings of Po/So phases from this region on the hydrophone installation at Wake Island. Of the events recorded by the OBH array in the Marianas through Kuril portion of the circum-Pacific arc, the largest (a 6.6 mb; event 40 in Table 2) occurred in the target area while all of the instruments were recording.

2. Po and So Propagation Velocities

Analysis of Po/So first arrivals to determine their propagation velocity was limited to those events with focal depths of 100 km or less that had travel paths

Fig. 2. Po and So first arrivals from a 6.6 mb event in the Kuril Islands (event 40, Table 2) shown in a reduced record-section form. The WIA hydrophone trace is not shown, but it would be at approximately the position of OBH 4. All traces were bandpass filtered between 4 and 10 Hz. Arrows indicate positions of first arrival picks.

to the OBH array lying entirely within the Northwestern Pacific Basin. In addition, for each type of phase (i.e., Po or So), an event was used only if there were clear first arrivals of that phase on at least two hydrophones in the array. First arrivals were picked independently by each author, and an arrival time was used only when the picks were in approximate agreement to a few tenths of a second. A total of thirty-six Po first arrivals from eight events, and twenty-nine So arrivals from a slightly different set of eight events were analyzed (see Table 2). Figure 2 shows the set of Po and So first arrivals from the 6.6 mb Kuril Islands event in

Fig. 3. Reduced travel-time plots of P_0 and S_0 first arrival data. The upper half shows the raw epicentral distance, travel-time (X, T) data, and the lower half shows the modified (X^*, T^*) data used to compute velocities across the array. Least-squares lines (solid lines) have been fit to each set of data and their respective equations are indicated. The dashed line on each upper plot is the same as the solid line on the corresponding lower plot, and vice versa. Event numbers are those given in Table 2.

reduced record section form. Arrows indicate the actual positions of first arrival picks.

A standard approach for determining the constant propagation velocity in a waveguide from travel time data is to use least-squares linear regression directly on those data. That is, an observed travel time, T_i , is modelled as a function of its epicentral distance, X_i , by the following equation: $T_i = X_i/V + I + \varepsilon_i$; where V is the propagation velocity, I is the travel-time intercept (i.e., the travel time at zero epicentral distance), and ε_i is the residual error of that observation. V and I are chosen so that the sum of the squares of all the residual errors is minimized. This is the approach that has primarily been used on the single station data (e.g., WALKER, 1977b). When this method is applied to the data of this experiment (see Fig. 3—upper half), the following travel-time equations result: $T = X/(8.14 \pm 0.05 \text{ km/s}) + (0.71 \pm 2.16 \text{ s})$ for P_0 , and $T = X/(4.75 \pm 0.04 \text{ km/s}) + (8.91 \pm 5.25 \text{ s})$ for S_0 .

Unfortunately, this method has certain shortcomings which can limit its usefulness. For one, it is highly dependent on knowing the epicenter and origin time precisely in order to determine each X_i and T_i . Errors in these parameters translate directly into uncertainties in V and I . In addition, any near source effects on the travel time that are not common to all of the events represented will also contribute to uncertainty or error in V and I . But, differences in near source effects are to be

expected in most data sets due to variations in focal depth, and to variations in the geometry of subduction in the source region, especially as it relates to the direction of the travel path. Also, the method assumes that the waveguide velocity is constant for all travel paths, or at least that variations in velocity are not correlated with epicentral distance. This may not be the case in the Northwestern Pacific, however, since for most previous data as well as the current OBH data set shorter travel paths are from earthquakes in the Marianas and Bonin Islands, while longer travel paths are more to the north from earthquakes in Japan and the Kuril Islands. Thus, any regional or azimuthal variation in the waveguide velocity can translate into additional scatter and even error in the determinations of V and I .

Data from the OBH array, combined with a slightly different least-squares model for determining V and I provided the opportunity to eliminate or greatly reduce these short comings. The form of this model is: $T_{ij} + \bar{T} - \bar{T}_i = (X_{ij} + \bar{X} - \bar{X}_i) / V + I + \varepsilon_{ij}$; where T_{ij} and X_{ij} are, respectively, the travel-time and epicentral distance from the i th earthquake to the j th hydrophone, $\bar{T}_i$ and $\bar{X}_i$ are, respectively, the mean travel-time and mean epicentral distance for observations of the i th event, $\bar{T}$ and $\bar{X}$ are the mean travel-time and mean epicentral distance for all observations, V is the propagation velocity, I is the travel-time intercept, and ε_{ij} is the residual error. Although the model looks complicated, if we let $T_{ij}^* = T_{ij} + \bar{T} - \bar{T}_i$ be the modified travel-time and $X_{ij}^* = X_{ij} + \bar{X} - \bar{X}_i$ be the modified epicentral distance, then the model becomes $T_{ij}^* = X_{ij}^* / V + I + \varepsilon_{ij}$ —which looks very much like the old model. As before, V and I are found by minimizing the sum of the squares of the residual errors. This is simply a least-squares line fit to the modified data.

This model has several advantages. For one, an error in travel-time that is common to all arrivals from a given event will only affect any corresponding T_{ij}^* to the extent that it modifies $\bar{T}$. That is because the error is cancelled out in the computation of T_{ij}^* by subtraction of the $\bar{T}_i$ term. And since the $\bar{T}$ term is constant and is common to all the T_{ij}^* terms in the analysis, errors of this type do not affect V , but only affect I . Thus, errors in travel-time that are due to errors in reported origin-times, variations in focal-depth, or variations in the subduction geometry have no effect on V . By a similar argument, errors in epicentral distance that are common to a given event, such as those produced by errors in the reported epicenter, also have no effect on V . And finally, any regional or azimuthal variations in the propagation velocity that happen to be correlated with epicentral distance will not affect V in a biasing way, but will simply be averaged together. This is because the average modified travel-time, $\bar{T}^*$, and average modified epicentral distance, $\bar{X}^*$, are constant and equal to $\bar{T}$ and $\bar{X}$ for all events. Essentially, this model is designed to measure Po and So propagation velocities across the array versus measuring the average propagation velocity over a portion of the Po or So travel-time curve as the other model is designed to do.

The lower half of Fig. 3 shows reduced travel-time plots of the modified Po/So data, with corresponding least-squares lines drawn through them. The equations for these lines are: $T^* = X^* / (7.96 \pm 0.05 \text{ km/s}) - (7.21 \pm 2.40 \text{ s})$ for Po, and $T^* = X^* /$

$(4.58 \pm 0.06 \text{ km/s}) - (12.84 \pm 7.61 \text{ s})$ for So. The values found for V and I are somewhat surprising. It is easy to see in the figure that the lines fit to these modified data are not well-suited to the unmodified data, and vice versa. In addition, these velocities are significantly lower than velocities of 8.33 km/s for Po and 4.77 km/s for So that have previously been reported for the northwestern Pacific (WALKER, 1977b). Also, the large negative intercepts appear to be somewhat hard to explain. They do imply that the Po/So energy must be propagating at a faster rate somewhere nearer the source than where these observations were made. Before discussing further the possible implications of these results in terms of Po/So propagation, it is worthwhile to examine the remaining sources of experimental error in the data that could be a cause of bias in these results.

3. Sources of Error

Possible sources of error include instrument-specific error such as OBH clock errors, OBH location errors, and errors caused by differences in crustal structure across the array. We estimate the magnitude of each of these types of errors to be no more than a few tenths of a second. Since the array is linear, and since all of the signals studied transversed the array in the same sense (i.e., from northwest to southeast), then any trend in these errors across the array would produce an error in V (and in I). For example, at 8 km/s, a trend amounting to one second across 500 km of the array would produce an error in V of 0.13 km/s (and an error in I of 5.7 s). Similarly at 4.6 km/s, a trend amounting to one second across 500 km would produce an error in V of 0.04 km/s (and an error in I of 5.4 s). Although there is no way to measure the actual magnitude of this effect in our small data set, precisely because it is inseparable from the velocities we are trying to measure, it is important to recognize this as a source of possible bias in the results.

Another important source of error which could bias V is the error in times of first-arrival picks. Both Po and So are emergent phases. Even when their coda have very large signal-to-noise ratios, their onsets often creep out of the noise. This characteristic, combined with the fact that signal-to-noise ratios decrease with increasing epicentral distance across the array, can easily produce a bias in the determination of V . The sense of this bias is negative; that is, its effect will reduce V by some amount. When looking at the Po/So first arrivals in Fig. 2, for example, it is easy to imagine that arrivals at greater epicentral distances might be picked at a later point in their emergent onset relative to arrivals at shorter epicentral distances. The magnitude of this pick bias might be a second or more across the array, and thus could have a significant effect on V and I , as was previously demonstrated.

Another source of error might be off-axis effects. With the exception of events 40 and 50, from the Kuril Islands, all events analyzed have Po/So signals that propagate across the array in a direction somewhat oblique to the axis of the array. Consequently, for a given event, the signal at each instrument has sampled, laterally, a slightly different part of the lithosphere. Thus, the velocity measured

between instruments could have errors caused by lateral differences between travel paths. If there is a trend to these errors, it will produce an error in V .

4. *Propagation Models*

In spite of the many possible causes for error in these determinations of V and I , it is interesting to speculate on models for Po/So propagation that could produce the apparently low propagation velocities across the array and the corresponding negative travel-time intercepts. BUTLER (1985) studied this same data set and suggested that azimuthal anisotropy in the lithosphere could explain the observations. Since the array is nearly parallel to magnetic lineations near Wake, propagation down the array could be in the slow direction, hence the relatively slow observed velocities. Furthermore, magnetic lineations in the source region near Japan and the Kuril Islands have a different orientation and are nearly perpendicular to the travel paths. Thus, the Po/So energy could propagate faster near the source—an effect that could produce negative travel-time intercepts. Unfortunately, attempts to measure the actual level of azimuthal anisotropy from these data are inconclusive, primarily because the linear array is unsuited for such measurements. More work needs to be done to assess this phenomenon and its effects.

A second model that might explain these observations is suggested by the apparent direct relationship between Po/So propagation velocities and lithospheric age (e.g., HART and PRESS, 1973; MCCREERY, 1981). In the Northwestern Pacific, the age of the lithosphere is generally increasing towards the subducting edge. Thus, Po/So energy propagating towards the OBH array might be systematically changing its velocity from faster to slower rates along the travel path. This would produce both slower velocities across the array, and negative travel-time intercepts. It is worthwhile to note that the average propagation velocity to the array that would be necessary to achieve a zero intercept is 8.13 km/s for Po and 4.68 km/s for So. These values are within the range of possible effects that might be produced by an increasing velocity due to lithospheric age.

5. *Comparisons to Older Data*

Figure 4 is a plot of most of the Po/So first arrival data collected by HIG in the Northwestern Pacific Basin since 1963. Much of this data is from WALKER (1977 b). Some modifications were made to a few of those published data (i.e., 13 Po velocities and 5 So velocities) because of errors that were found in some of the original hydrophone coordinates. The formerly secret status of those hydrophone sites, and the circuitous route by which coordinates were originally obtained are the likely cause for those unfortunate errors. Recomputation of epicentral distances for the erroneous data produced two significant changes to the results of that report. First, it eliminated an apparent discontinuity in propagation velocity which had been suggested at about 30° epicentral distance. Second, it lowered the Po/So

Fig. 4. Average velocity (X/T) versus epicentral distance (X) for P_0 and S_0 with Northwestern Pacific travel paths and focal depths of 100 km or less. Included in this data set are island seismograph and deep ocean hydrophone data collected before the OBH experiment (dots), data from WIA collected between 1978 and 1980 (open triangles), and data from the OBH experiment (solid triangles). Superimposed on these data are curves representing the travel-time line computed from the modified OBH data, with plus and minus one standard deviation.

propagation velocities somewhat, a result that moved those values nearer to the values found using the modified OBH array data.

First arrival data are plotted in Fig. 4 as average velocity (X/T) versus epicentral distance (X). Superimposed on this plot are curves representing the P_0 and S_0 travel-time lines computed from the modified OBH array data (lower half of Fig. 3), with plus and minus one standard deviation. Standard deviations were computed not only from the error in the V and I values of the travel-time line, but also from an additional assumed error of 10 km in epicentral distance and 1 s in origin time, since these additional types of error are present in the X/T data. In spite of the fact that the travel-time lines do not fit the unmodified OBH array data very well, as illustrated in Fig. 3, they do seem to fit somewhat better to this larger collection of Northwestern Pacific data. The most prominent exceptions to the fit are at epicentral distances less than about 12° , where average velocities are much lower, and in S_0 at distances between about 12° and 27° , where most data fall below the curve.

The obvious discontinuity in these data at about 12° epicentral distance requires explanation. It could be the manifestation of a different branch of the P_0

and So travel-time curves. The average X/T for the data at less than 12° is 7.96 ± 0.16 for Po and 4.58 ± 0.05 for So. Except for the standard deviations, these values are identical to the propagation velocities found across the OBH array. Qualitatively then, a model for these observations might be that the Po/So first arrival energy travels entirely within the waveguide out to distances of 12° ; but beyond 12° the first arrival energy has propagated along another slightly faster path before being trapped in the waveguide. However, another explanation might simply be that these data at less than 12° are incompatible with the rest, since 10 out of the 12 data points for both Po and So are from only one seismic station—Marcus Island. These data could be indicators of azimuthal anisotropy, since their associated travel paths to Marcus appear to be primarily in the slow direction based on nearby magnetic lineations. Additional data at distances near 12° will be needed to resolve the questions pertaining to this apparent discontinuity.

In this regard, we note that three additional data points are available from OUCHI *et al.* (1983) for distances less than 12° . These observations were acquired during an OBS experiment west of Marcus Island in June and July of 1981. The earthquake which provided the data, recorded at distances of approximately 9.4° , 10.5° , and 11.5° , occurred off the east coast of northern Honshu. The average velocities (X/T) estimated from their reduced travel time curves are approximately 7.8 km/s. Although these values are comparable to those found at Marcus Island, additional data is still needed for distances less than 12° . We note that OUCHI *et al.* (1983) also provide a plausible explanation for some of the unusually high Po velocity values observed in the Western Pacific. They believe that low-frequency (i.e., 3 Hz), high-velocity (i.e., ≥ 8.39 km/s) arrivals apparent at distances from about 9 – 15° are probably mantle-refracted P phases. Furthermore, they believe that such phases could be misinterpreted as Po arrivals.

6. Conclusions

Data collected by a 1,500-km OBH array deployed for two months in the Northwestern Pacific Basin near Wake Island have provided some new information about the propagation of Po/So phases. With this array, propagation velocities of first arrivals from shallow-focus (<100 km) events have been measured by a different and hopefully more accurate method than was previously possible with single station data. The velocities found are 7.96 ± 0.05 km/s and 4.58 ± 0.06 km/s for Po and So, respectively, measured at epicentral distances between 20° (2,200 km) and 34° (3,800 km). Associated with these propagation velocities in the linear travel-time equation are intercepts equal to -7.21 ± 2.40 s and -12.84 ± 7.61 s for Po and So, respectively. These large negative intercepts imply higher propagation velocities over some part of the travel path nearer to the source than where the data were recorded. Higher velocities might be achieved by the effects of azimuthal anisotropy in the waveguide or by increasing propagation velocity with lithospheric age. Travel times of Po/So first arrivals from older data collected by HIG since 1963, for

shallow-focus events with Northwestern Pacific Basin travel paths greater than 12° epicentral distance, are modeled reasonably well by the travel-time line reported here. However, data at epicentral distances less than 12° (also collected prior to this experiment) have much lower average velocities (X/T) and may follow a different Po/So travel-time branch than the one described by the OBH data. Additional data is needed to resolve this discontinuity at 12° .

Regarding attenuation, we note that the magnitude 6.6, Kuril Island event (event 40, Table 2) is especially important. Its large magnitude and its location in the area targeted by the array produced Po/So phases that travelled down the axis of the 1,500-km deployment line with sufficient energy to be observed with high signal/noise ratios on many of the instruments. Using this data BUTLER *et al.* (1987) found frequency dependent Q 's for both Po and So. Q values for Po ranged from about 300 at 2.5 Hz to 1,500 at 17.5 Hz, and values for So ranged from about 400 at 2.5 Hz to 3,000 at 22.5 Hz.

In conclusion, it is our hope that all of these results will provide meaningful constraints on modeling efforts, resulting ultimately in a better understanding of the oceanic lithosphere and the dynamics of sea-floor spreading.

This research was supported by the Office of Naval Research (code 425 GG). Funds for support of the hydrophone station at Wake Island were provided primarily by the Air Force Office of Scientific Research with supplementary support from the U.S. Arms Control and Disarmament Agency. The authors express thanks to Dave Byrne, Grant Blackinton, Bob Mitiguy, Dave Barrett, and Fred Duennebieer for their help in modifying, testing, launching, and recovering the OBH's. Appreciation is also expressed to Al David and Kentron Corporation for their part in maintaining the recording station at Wake. The authors also thank George Sutton, Neil Frazer, and Rhett Butler for reviewing earlier versions of this manuscript, and Rita Pujale for editorial assistance. Hawaii Institute of Geophysics Contribution No. 1903.

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